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## ORIGINAL PAPER

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## Mechanical aspects of sedimentary basin formation: development of integrated models for lithospheric and surface processes

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**Abstract** Different assumptions for the thermo-mechanical properties of the lithosphere strongly affect predictions inferred from quantitative sedimentary basin modeling. Examples from various basins, selected as natural laboratories, illustrate the importance of incorporating a finite strength of the extending lithosphere in forward stratigraphic modeling of large-scale basin stratigraphy. Current models can effectively couple erosion at uplifted rift shoulders of extensional basins with the basin fill architecture of the subsiding basin compartments. Modeling of the synrift strata integrates spatial scales characteristic for subbasins, such as the Oseberg field in the North Sea, with large-scale lithospheric properties characterizing the bulk strength of extending lithosphere. Modeling of compressional basins in foreland fold-and-thrust-belt settings can effectively link lithospheric flexure with surface processes. Scales pertinent to short-term spatial and temporal variations in basin fill and basin deformation can now be addressed, allowing the quantitative investigation of consequences of different modes of thrusting for basin fill geometry and facies characteristics.

**Key words** Sedimentary basin modeling · Quantitative stratigraphy · Lithospheric dynamics · Thermo-mechanical models · Extensional tectonics · Rift shoulders · Compressional basins · Fold-and-thrust belts · Erosion · Surface processes

### Introduction

Since the advent of plate tectonics, the power of the sedimentary basin record for extracting information on the temporal evolution of the underlying continental lithosphere has been increasingly recognized (e.g., Zieg-

ler 1990, 1996; Cloetingh et al. 1993, 1995a, b, 1996; Cloetingh, Sassi and Task Force Team 1994). At the same time, it has also become quite manifest that for a full understanding of the interplay of tectonics and climate and their controls on the sedimentary record a good understanding of the role of the lithospheric processes involved is a prerequisite. This is obvious, considering the importance of topography for climate and the feedback of these factors with erosion and sediment supply (Schlager 1993). More recently, also the potential importance of feedback between apparently independent processes, such as lower crustal flow and surface erosion (Burov and Cloetingh 1997), has been identified as an example requiring an integration of lithospheric and near-surface processes. Basin modeling, as it stands now, has great potential to provide the starting point for such an integrated approach, especially if the modeling is constrained by high-quality data sets. Sequence stratigraphy (Vail et al. 1977; Haq et al. 1987; Posamentier and Allen 1993) and the availability of high-quality seismic data from programs such as ECORS, BIRPS, DKORP, and EUROPROBE on the crust underlying basins and the basin fill, as well as extensive data sets collected as part of the industrial search for hydrocarbons (Ziegler 1990; Ziegler and Roure 1996), has revolutionized basin research and form the foundation for the development of new modeling concepts and tools. In this context the constraints provided by 3D seismic data on the fine structure of basin deformation illustrate the necessity to expand the role of structural geology in basin modeling, linking the different spatial and temporal scales involved in both hydrocarbon exploration and production aspects of basin research (Gabrielsen and Strandenes 1994).

### The first generation of basin models

After the realization that subsidence patterns of Atlantic-type margins, corrected for effects of sediment load-

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ing and paleobathymetry, displayed the typical time-dependent decay characteristic of ocean-floor cooling (Sleep 1971), a large number of studies was undertaken aimed at restoring the quantitative subsidence history of basins on the basis of well data and outcropping sedimentary sections. With the introduction of geohistory analysis approaches (Van Hinte 1978) and backstripping analyses algorithms (Steckler and Watts 1981; Bond and Kominz 1984), the late 1970s and early 1980s marked a phase where basin analysis essentially stood for backward modeling namely reconstructing from data the tectonic subsidence. These quantitative subsidence histories provided constraints on which development of forward, conceptually driven basin formation modeling could build. For extensional basins this development started in the late 1970s, with the appreciation of the importance of the lithospheric thinning and stretching concepts in basin formation (Salveson 1976, 1978). After initiation of mathematical formulations of stretching concepts (McKenzie 1978) in forward extensional basin modeling, a large number of basin fill simulators appeared on the market. These packages focused on the interplay of thermal subsidence, sediment loading, and eustatic sea-level changes. Smooth post-rift subsidence behavior was modulated by changes in sediment supply and eustatic sea level to arrive at commonly observed more episodic and irregular subsidence patterns. Another approach, often taken in existing packages, is to input a subsidence curve, making the basin modeling package essentially a tool to fill in an adopted accommodation space (Burton et al. 1987; Lawrence et al. 1990). This approach separates the post-rift evolution of an extensional basin from the tectonics of basin formation, equating tectonic subsidence during the post-rift phase, "being by its nature slow," to the superposition of thermal subsidence and lithospheric flexure (Watts et al. 1982).

A similar set of assumptions was made to describe the synrift phase. In the most simple version of the stretching model (McKenzie 1978) thinning was described resulting from a more or less instantaneous extension. In these models a component of lithosphere mechanics was obviously lacking. On a smaller scale tilted fault block models entered the market and were used for modeling of basin fill on the scale of half-graben models. Such models essentially decouple the response of the brittle upper crust from the deeper lithospheric levels during rifting phases (e.g., see Kusznir et al. 1991; Waltham 1992).

A noteworthy feature of most of the existing software packages is their emphasis on the basin subsidence record and their very limited capability to handle uplift and differential subsidence and uplift patterns in a process-oriented, internally consistent manner (e.g., see Larsen et al. 1992; Dore et al. 1993).

To a large extent the same is true for most of the software packages available for modeling of compressional basins. The realization of the importance of the lithospheric flexure concept, relating topographic load-

ing of the crust by an overriding mountain chain to the generation of accommodation space, was recognized as early as 1973 by Price in his paper on the Canadian Rocky Mountains foreland fold-and-thrust belt (Price 1973). Also, here it took several years before quantitative approaches started to develop, investigating the effects of lithospheric flexure on foreland basin stratigraphy (Beaumont 1981). The success of the flexural basin stratigraphy modeling, capable of incorporating subsurface loads related to plate tectonic forces operating on plates (e.g., Van der Beek and Cloetingh 1992; Peper et al. 1994), led relatively recently to the realization that there was a need to incorporate more structural complexity in these models.

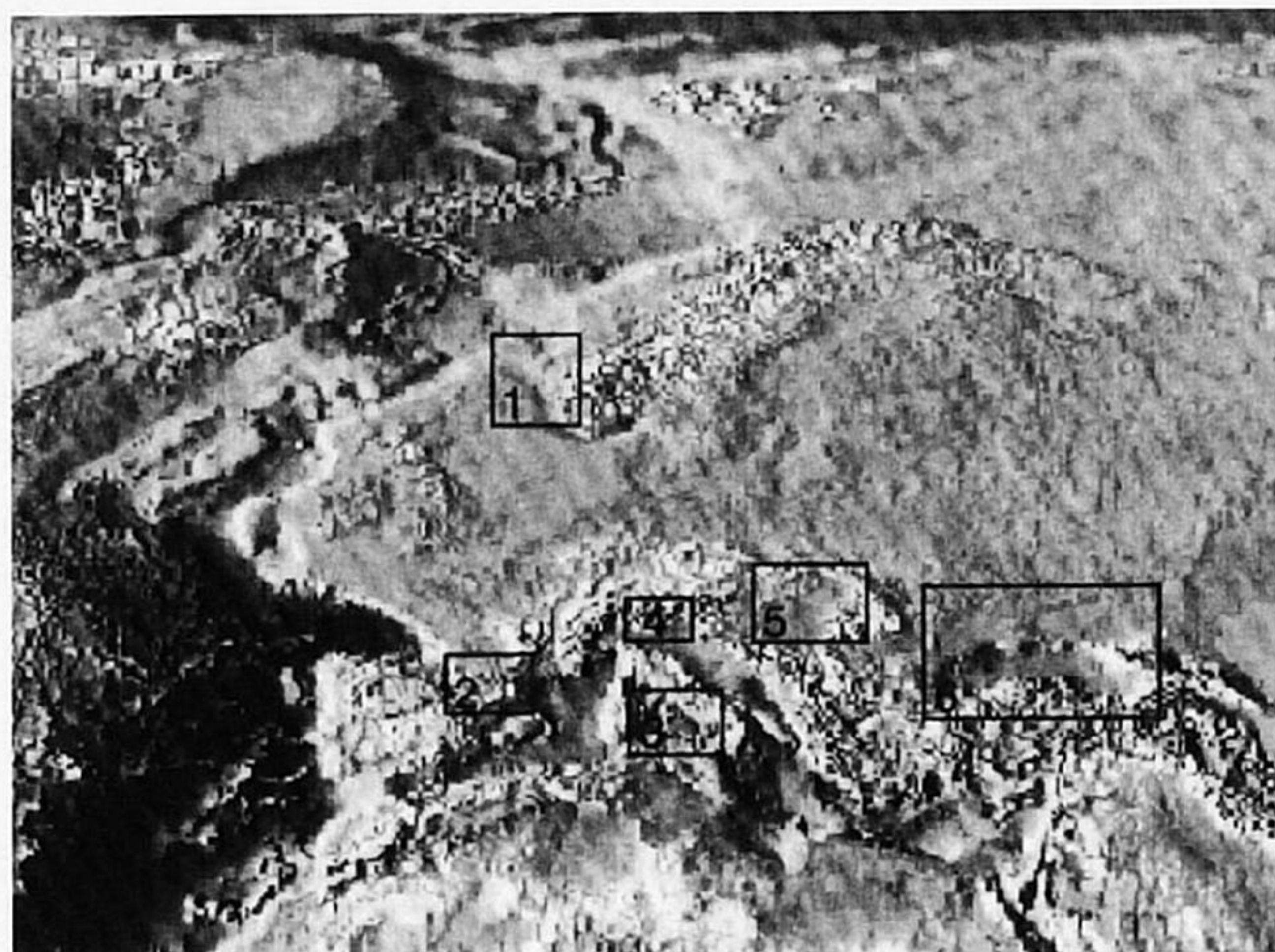
This is not surprising, considering the strong emphasis which was placed by commercial packages on the development of successful tools for the simulation of thermal maturation and fluid-flow migration (e.g., see Parnell 1994). In this respect it must be realized that in the academic world the necessary knowledge on lithospheric mechanics and basin deformation has been developed in the past few years. This was only possible after a bridge was established between researchers studying deeper lithospheric processes and those who analyzed the record of vertical motions, sedimentation, and erosion in basins. This permitted the development of basin analysis models integrating structural geology and lithosphere tectonics.

The focus of modeling activities in joint European projects, such as the Integrated Basin Studies project (IBS) carried out by the Amsterdam Tectonics group, was on the quantification of mechanical coupling of lithosphere processes to the near-surface expression of tectonic controls on basin fill (Cloetingh et al. 1995a, b, 1996). This invoked a process-oriented approach, linking different spatial and temporal scales in the basin record. An essential part of our modeling philosophy was the testing and validation in close partnership with other academic and industrial research groups, of modeling predictions in well-constrained natural laboratories (see Cloetingh, Sassi and Task Force Team 1994). Basins were selected on the basis of the availability of high-quality databases for both deeper crustal levels (deep seismic reflection and refraction) and basin fill (reflection seismic, wells, and outcrops). Figure 1 shows the location of a number of the selected extensional basins. Our strategy was to develop on the basis of these data sets new modeling components. Applying the same modeling approach to different basins provided an opportunity to compare key parameters for basin evolution, shedding light on the controls underlying observed similarities and differences in basin histories.

### **Control of large-scale rheology on basin shape**

Foreland basins owe their existence to the capacity of the lithosphere to support topographic loads, such as



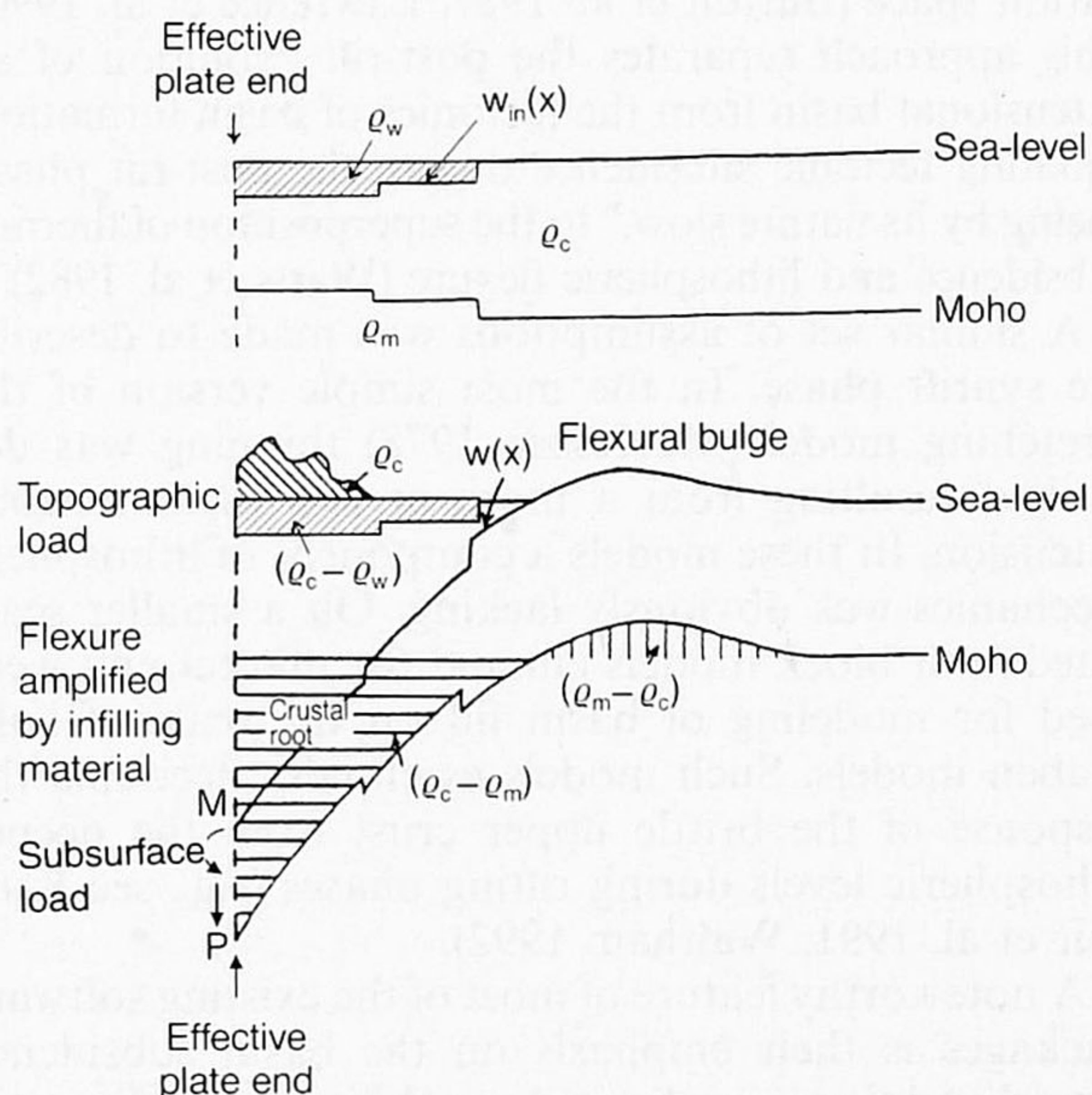


**Fig. 1** Location of the basins discussed in this text. Numbers refer to the following basins: 1 Oseberg area (northern Viking graben); 2 Pyrenees and Ebro basin; 3 Tyrrhenian Sea; 4 southern Alps; 5 Pannonian basin; 6 Black Sea

mountain chains, and to bend down over areas often exceeding the spatial scale of the load. The width of the resulting depression, the foreland basin, provides an indication of the mechanical strength of the underlying lithosphere. A zero-strength lithosphere would simply sink under the load, without the creation of a foreland basin accompanying this process (Airy isostasy). In contrast, a strong lithosphere, characteristic of typical cratonic settings, would permit the development of wide and relatively shallow foreland basins (flexural isostasy). Examples of such basins are associated with the Canadian Rockies and the Appalachian fold-and-thrust belt of North America and their Precambrian cratonic foreland crust (e.g., Beaumont 1981; Quinlan and Beaumont 1984). The Alpine foreland basins, so characteristic of the geology of Europe, developed on younger lithosphere that was stratified after the Variscan orogeny and subsequently was affected by Mesozoic rifting activity. The time lapse between rifting and synorogenic flexural deformation of the foreland lithosphere is very variable (Pyrenees: ca. 30 Ma; Apennines: ca. 130 Ma). The relatively modest widths of the foreland basins of the Carpathians (Zoetemeijer and Cloetingh 1997; Matenco et al. 1997), Pyrenees (Millan et al. 1995), the Betics (Van der Beek and Cloetingh 1992), the Apennines (Zoetemeijer et al. 1993), and the eastern Alps (Andeweg and Cloetingh 1997) reflect the presence of relatively weak lithosphere. In addition to the role of topographic loading by mountain chains and the sedimentary loading of the foreland basins, additional forces operating on the lithosphere, such as slab pull, slab detachment, and slab roll back, play an important role in shaping the basin geometry (e.g., Millan et al. 1995).

In the first generation of flexural foreland basin models, the different loading components and forces were

represented by a system of vertical loads, shear forces, and bending moments (Royden 1988, 1993). This approach is illustrated in Fig. 2. A characteristic feature of these models is the negligence of horizontal stresses, which have been shown to play a potentially important role in the geometry of the basin and the architecture of the basin fill (Peper et al. 1994). These models were utilized to extract information on the mechanical properties of the lithosphere, which was treated as an elastic plate rather than as a depth-dependent rheologically stratified beam. Consequently, the models yielded rough estimates for the effective elastic thickness of the lithosphere and a first approximation for the bulk rheology of the continents. This approach permitted examination of the role of time elapsed since the last thermal perturbation of the lithosphere (generally a rift phase) on the deflection during the subsequent foreland basin phase. In this respect, Desegaulx et al. (1991) have explained the rapid Cenozoic subsidence of the Aquitaine foreland basin (SW France) in terms of two superimposed effects resulting from the pre-orogenic rifting phase. These authors have shown that the foreland flexure was affected by the weakening effects in the thinned lithosphere resulting from the pre-orogenic extension and an associated thermal perturbation. The subsequent cooling of the extended lithosphere has caused an additional subsidence equivalent to an extra tectonic subload in the



**Fig. 2** Concept of lithospheric flexure in foreland basin evolution. The regional deflection of the bending lithosphere is the combined result of topographic loading by the overlying mountain chain, the weight of the sediments in the foreland basin, and plate tectonic forces operating on the underthrust lithosphere. The shape of the flexural depression is affected by the interplay of lithospheric strength and the magnitude of these differential loads. (After Royden 1993)

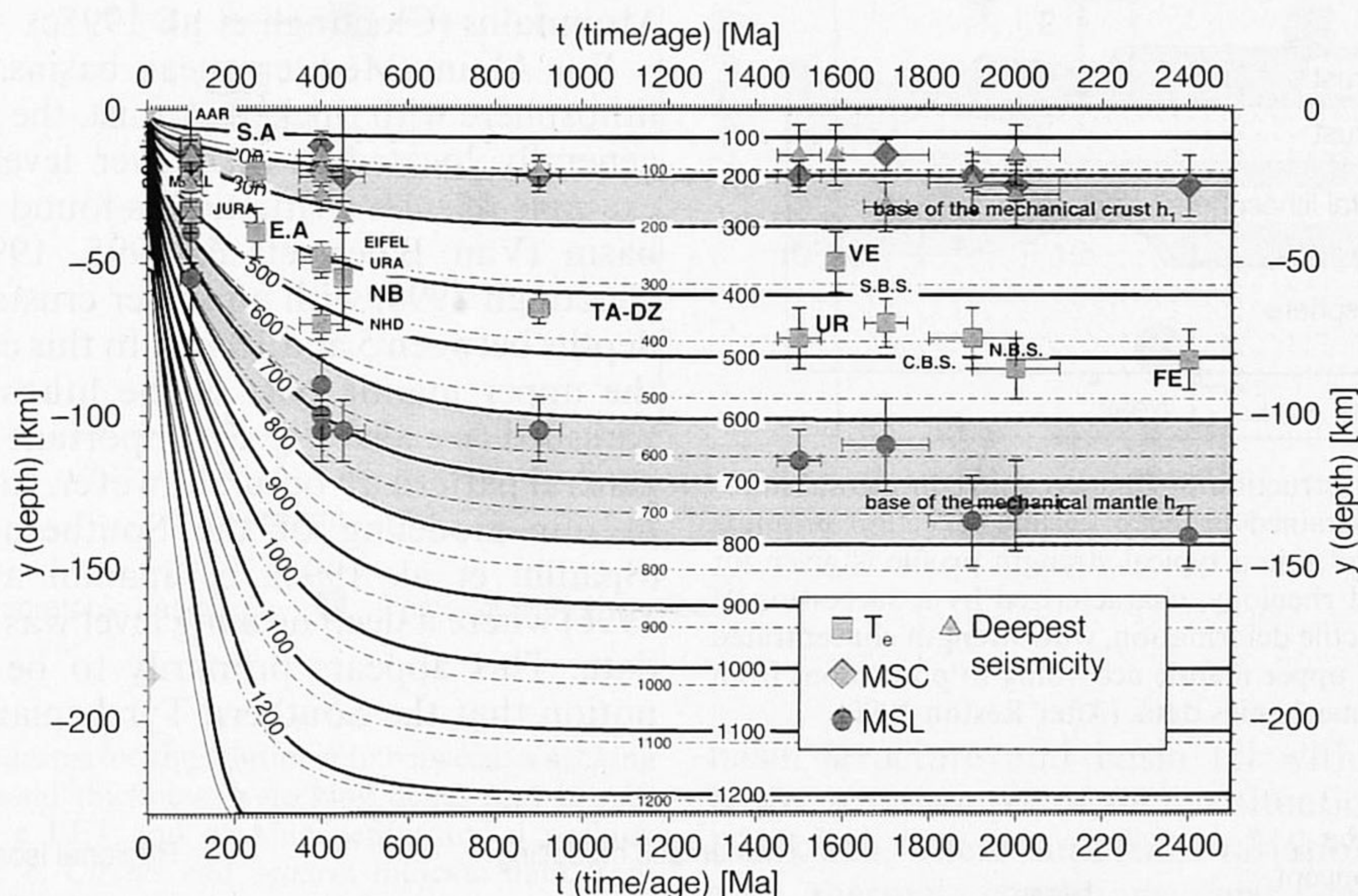


foreland basin phase of the Aquitaine basin (Desegaulx et al. 1991).

The concept of thermomechanical age provided the framework for effective elastic thickness (EET) estimates. Figure 3, which was constructed on the basis of a large data set for Eurasian foreland basins (Cloetingh and Burov 1996), illustrates a general trend of increasing elastic plate thickness with thermo-mechanical age. Also plotted in Fig. 3 are the predictions for the bulk rheology of the lithosphere based on extrapolations from rock-mechanics data, constrained by crustal geophysical data and thermal models. A characteristic feature of these models is the incorporation of a quartz-dominated upper crustal rheology and an olivine controlled mantle rheology (see also Fig. 4). These models are cast in terms of the depth to the bottom of the mechanically strong upper part of the crust (MSC) and the mechanically strong part of the upper mantle lithosphere (MSL). Analysis of Fig. 3 demonstrates that the mechanical properties of the crust are not affected very strongly by cooling with increasing age. In marked contrast is, however, the age-dependent predicted increase in the thickness and strength of the mantle component of the continental lithosphere. The band of

EET values, representing the integrated average response of the lithosphere to surface loads, reflects the combined response of depth-dependent rheological models including two mechanically strong layers, corresponding to the upper crust and the upper parts of the mantle lithosphere, separated by the mechanically weak lower crust. The degree of coupling and decoupling between these two layers and the scatter of the data reflects to a large extent the importance of mechanical weakening by horizontal stresses in the lithosphere (Cloetingh and Burov 1996).

The notion of a possible zone of decoupling between a strong upper crust and a strong upper mantle is also important in the context of extensional basin formation. Over the past decade vigorous discussions have been going on concerning the relative importance of pure shear (McKenzie 1978) and simple shear (Wernicke 1985) mode of extension. Figure 4 suggests that in the presence of a weak lower crustal layer, decoupling of the mechanically strong upper crust from the even stronger upper mantle, the zone and symmetry of upper crustal extension, does not necessarily have to coincide with the zone and symmetry of upper mantle extension. This is particularly the case if the upper crust is



**Fig. 3** Compilation of observed and predicted values of effective elastic thickness (EET) depth to bottom mechanically strong crust (MSC) and depth to bottom mechanically strong mantle lithosphere (MSL) against the age of the continental lithosphere at the time of loading and comparison with predictions from thermal models of the lithosphere. Depth on vertical axis, age on horizontal axis, isotherms as labeled contours. Isotherms marked by *solid lines* are for models that account for additional radiogenic heat production in the upper crust. *Dashed lines* correspond to pure cooling models for continental lithosphere. The equilibrium thermal thickness of the continental lithosphere is 250 km. *Shaded bands* correspond to depth intervals marking the mechanical base of the crust (MSC) and the mantle portion of the lithosphere (MSL). *Squares* correspond to EET estimates, *circles* indicate MSL estimates, and *diamonds* correspond

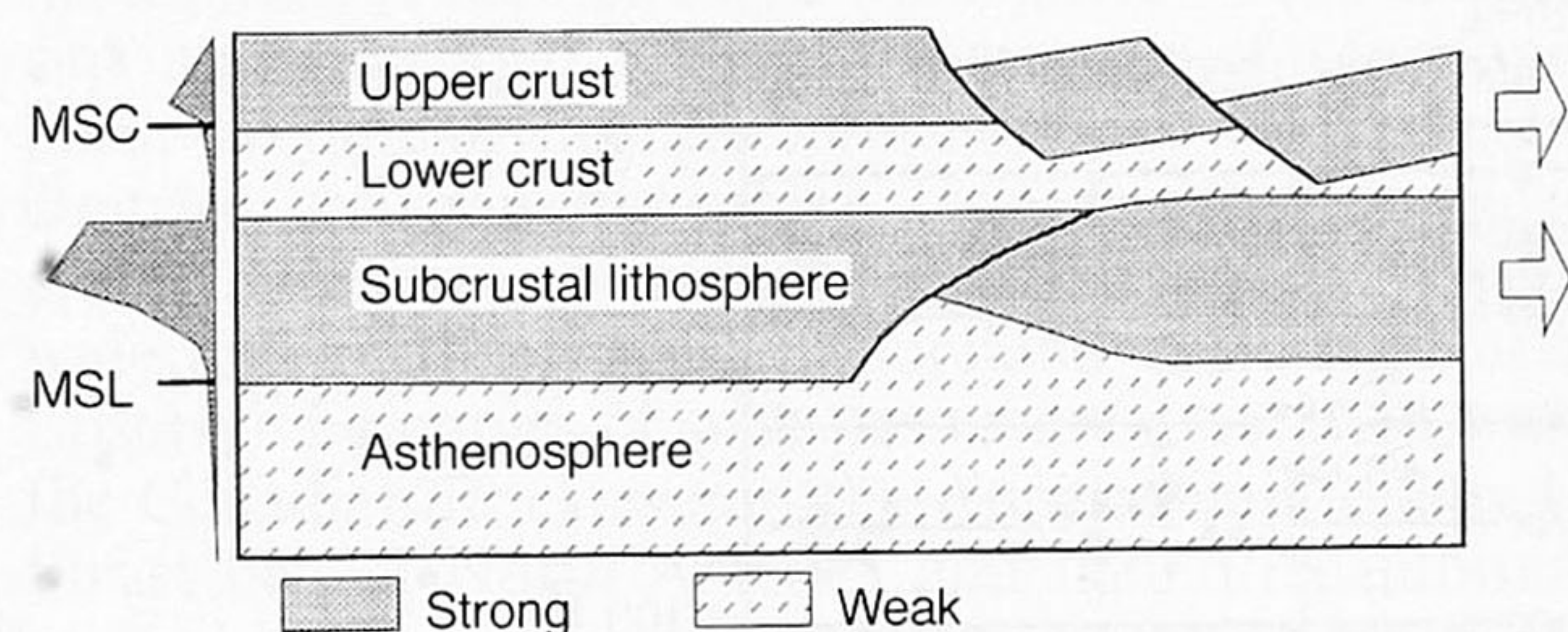
to estimates of MSC. *Bold letters* correspond to directly estimated EET values derived from flexural studies on, for example, foreland basins, *Thinner letters* indicate indirect rheological estimates derived from extrapolation of rock-mechanics studies. The data set includes I: Old thermomechanical ages (1000–2500 Ma): northernmost (N.B.S.), central (C.B.S.), and southernmost Baltic shield (S.B.S.); Fennoscandia (FE); Verkhoyansk plate (VE); Urals (UR); Carpathians; Caucasus, II: Intermediate thermomechanical ages (500–1000 Ma): North Baikal (NB); Tarim and Dzungaria (TA-DZ); Variscan of Europe: URA, NHD, EIFEL. III: Younger thermomechanical ages (0–500 Ma): Alpine belt: JURA, MOLL (Molasse), AAR; southern Alps (SA) and eastern Alps (EA); Ebro basin; Betic rifted margin; Betic Cordilleras. (After Cloetingh and Burov 1996)



weakened by pre-existing discontinuities favoring its simple shear extensional deformation (Ziegler 1996). A better appreciation of the role played by rheology during basin formation and the advent of corresponding modeling capabilities during the past few years has increasingly shifted the focus of attention away from these end members of lithospheric extension. At the same time it was increasingly realized that rheological decoupling of the upper crust and mantle lithosphere can play an important role in the structural style of intraplate compressional deformations (e.g., the Rocky Mountain and Bohemian massif; Ziegler et al. 1995).

### Finite strength of the lithosphere in extensional basin formation

An important step in our approach to extensional basin modeling has been the implementation of finite lithospheric strength. The standard stretching models built assume that during rifting the strength of the lithosphere is zero. This assumption is incorrect in light of what we know presently about lithospheric mechanics

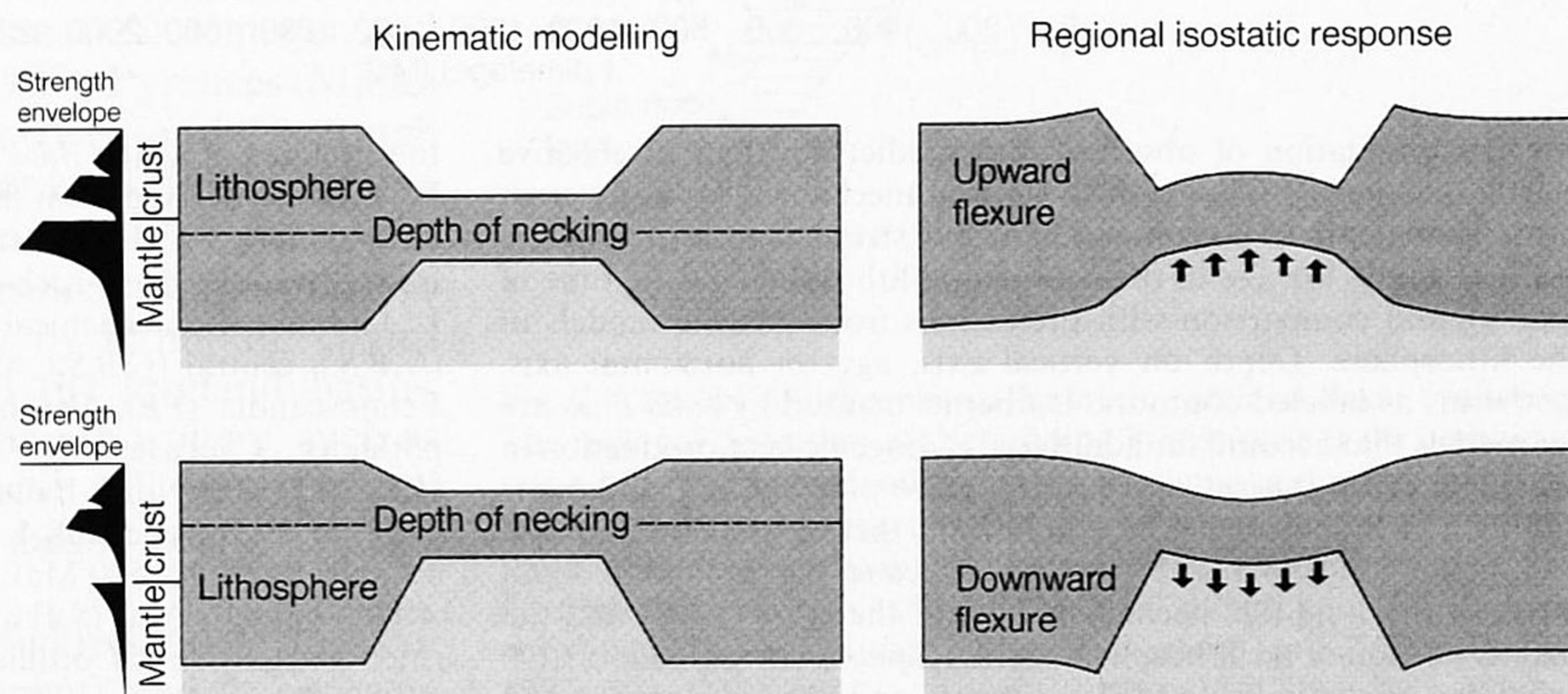


**Fig. 4** Kinematic reconstruction of Phanerozoic continental lithosphere extension, constrained by deep seismic reflection profiling studies. At the left-hand side a typical strength profile is given for a continental stratified rheology, characterized by a succession of zones of brittle and ductile deformation, with strength concentrated in the upper crust and upper mantle according to predictions from extrapolation of rock-mechanics data. (After Reston 1990)

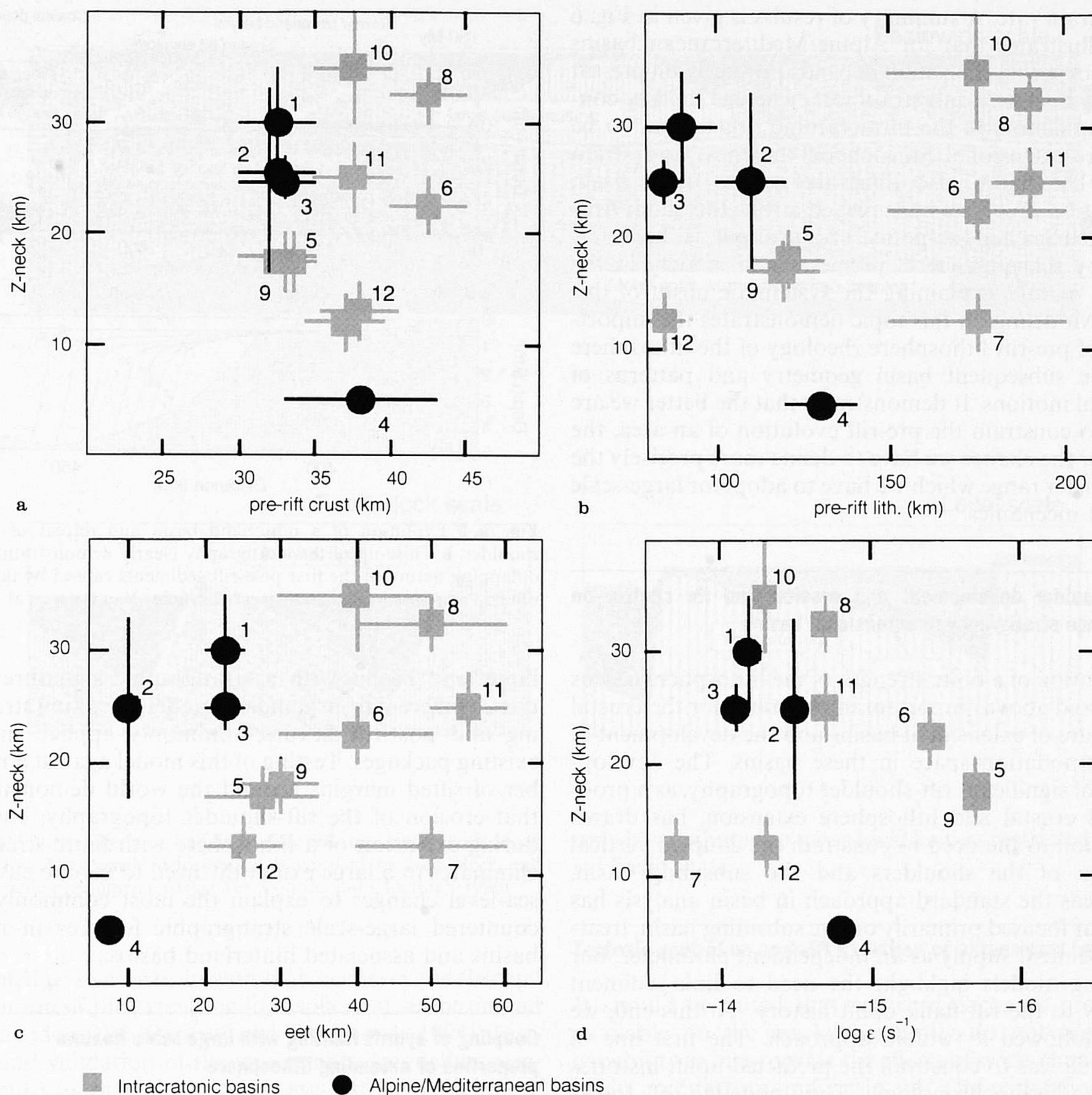
(e.g., Ranalli 1995; Burov and Diament 1995). Finite-element models have explored the large-scale implications of a finite lithospheric strength and in particular its sensitivity to the presence of fluids in the crust and mantle lithosphere (Braun and Beaumont 1989; Dunbar and Sawyer 1989; Govers and Wortel 1993; Bassi 1995). These dynamic models require intensive computing, are expensive to run, and thus are not suitable for an industrial user-oriented environment. However, they have provided the background for a more user-friendly class of kinematic models targeted on modeling of rift-shoulder uplift and basin fill (Cloetingh et al. 1995c). These kinematic models invoke the concept of lithospheric necking around one of its strong layers during extensional basin formation (see Braun and Beaumont 1989; Kooi et al. 1992; Spadini et al. 1995a). Figure 5 illustrates the basic features of these models and their relationship with the internal strength distribution of the lithosphere. In the presence of a strong layer in the subcrustal mantle, the level of lithospheric necking is deep, inducing pronounced rift-shoulder topography. This type of response is to be expected if extension affects cold and correspondingly strong intracontinental lithosphere; it is commonly observed in intracratonic rifts and rifted margin such as, for example, the Red Sea and the Transantarctic Mountains (Cloetingh et al. 1995c).

For Alpine/Mediterranean basins, located on weak lithosphere with thickened crust, the level of necking is generally located at shallower levels (Fig. 6). A key example of such a situation is found in the Pannonian basin (Van Balen et al. 1995, 1997; Horváth and Cloetingh 1996) with an upper crustal necking level at depths between 5 and 10 km. In this case the strength of the upper mantle part of the lithosphere has almost vanished (see also Fig. 5). Important exceptions to this general pattern do occur, however, such as encountered in our modeling of the Southern Tyrrhenian Sea (Spadini et al. 1995a, b; Spadini and Podladchikov 1996) where a deep necking level was required to fit the data. This appears primarily to be the result of the notion that the Southern Tyrrhenian basin developed

**Fig. 5** Illustration of the lithosphere necking concept with a finite flexural rigidity during extensional basin formation. An upward state of flexure is induced by a deep necking level, and a downward state of flexure is induced by a shallow necking level. (After Kooi et al. 1992)







**Fig. 6** Correlation diagrams for the relationship between: **a** necking depth and pre-rift crustal thickness; **b** necking depth and pre-rift lithosphere thickness; **c** EET and necking depth; and **d** necking depth and strain rate  $\epsilon$ . Circles and squares indicate data from Alpine/Mediterranean basins and intracratonic rifts, respectively. Numbers refer to the following basins: 1 Gulf of Lion; 2 Valencia trough; 3 southern Tyrrhenian Sea; 4 Pannonian basin; 5 North Sea; 6 Baikal rift; 7 Saudi Arabia Red Sea; 8 Transantarctic Mountains; 9 Barents Sea; 10 East African Rift; 11 western Black Sea; 12 eastern Black Sea. (After Cloetingh et al. 1995c)

largely on Hercynian crust with a significant component of the bulk strength of the lithosphere in its mantle component. The Tyrrhenian Sea basins has been subject to many detailed studies (for a recent review see Cloetingh et al. 1995c). In other areas this concept has led to a better understanding of lateral variations in

basin structure and basin fill within one and same basin. An example of such a situation was seen in the Black Sea, where important variations in necking level and thermal conditions, was deciphered by the modeling, between the eastern and western Black Sea, are related to differences in the timing and mode of basin formation (Spadini et al. 1996; Spadini and Cloetingh 1997; Robinson et al. 1995).

In a detailed comparative study of the basins investigated through the EU funded Integrated Basin Studies (IBS) project, integrating data from basins outside the IBS areas, and employing the same modeling technology (Cloetingh et al. 1995b, c), we were able to systematically investigate the dependence of necking depth on other parameters such as pre-rift crustal thickness, pre-rift lithospheric thickness, effective elastic thickness,



and strain rate. A summary of results is given in Fig. 6 and illustrates that for Alpine/Mediterranean basins the necking-level position depends primarily on pre-rift crustal thickness and strain rate, whereas the key controlling factors in the intracratonic rifts appear to be the pre-extensional lithospheric thickness and strain rate. The figure also illustrates where other basin formation processes have played a role: the Saudi Arabia–Red Sea margin (point 7), for example, is characterized by the presence of plume-related activity in the upper mantle, explaining the systematic misfit of this case. Modeling on this topic demonstrates the importance of pre-rift lithosphere rheology of the lithosphere on the subsequent basin geometry and patterns of vertical motions. It demonstrates that the better we are able to constrain the pre-rift evolution of an area, the greater the chance we have to define more precisely the parameter range which we have to adopt for large-scale synrift mechanics.

### Rift-shoulder development and erosion and its control on sequence stratigraphy in extensional basins

The notion of a finite strength of the lithosphere has, as discussed above, important implications for the crustal structure of extensional basins and the development of accommodation space in these basins. The development of significant rift-shoulder topography, as a product of crustal and lithosphere extension, has drawn attention to the need to constrain the coupled vertical motion of the shoulders and the subsiding basin. Whereas the standard approach in basin analysis has thus far focused primarily on the subsiding basin, treating sediment supply as an independent parameter, our necking models highlight the need to link sediment supply to the rift-flank uplift history. To this end, we have followed a twofold approach. The first line of research was to constrain the predicted uplift histories by geothermochronology. The modeling of track-length distributions enabled the use of fission-track data in an effort to backstack the eroded sediments from their present position in the basin to their source on the rift shoulder in an attempt to obtain a better reconstruction of rift-shoulder geometry. This work (e.g., Van der Beek et al. 1995; Rohrmann et al. 1995) has led to a better understanding of the timing and magnitude of rift shoulder uplift in, for example, the Norwegian margin, shedding light on the observed relationships between on shore uplift and the presence of thick late Cenozoic sedimentary wedges offshore. A second line of research focused on the development of a model for basin fill simulation, integrating the effects of rift-shoulder erosion through hill-slope transport and river incision with sediment deposition in the basin. As illustrated in Fig. 7 (see Van Balen et al. 1995), these models predict prograding sedimentary wedges within the extensional basin and the development of

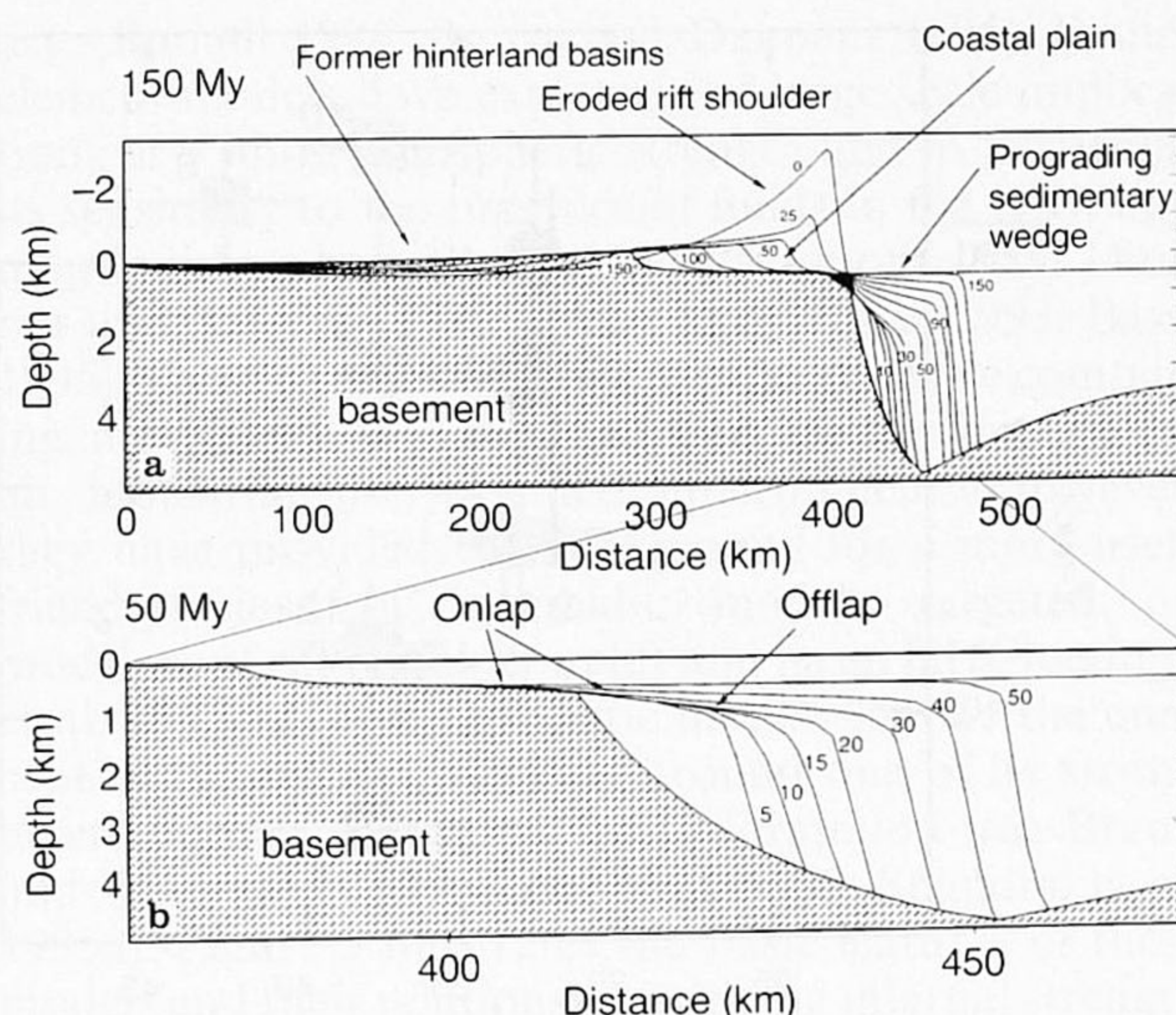


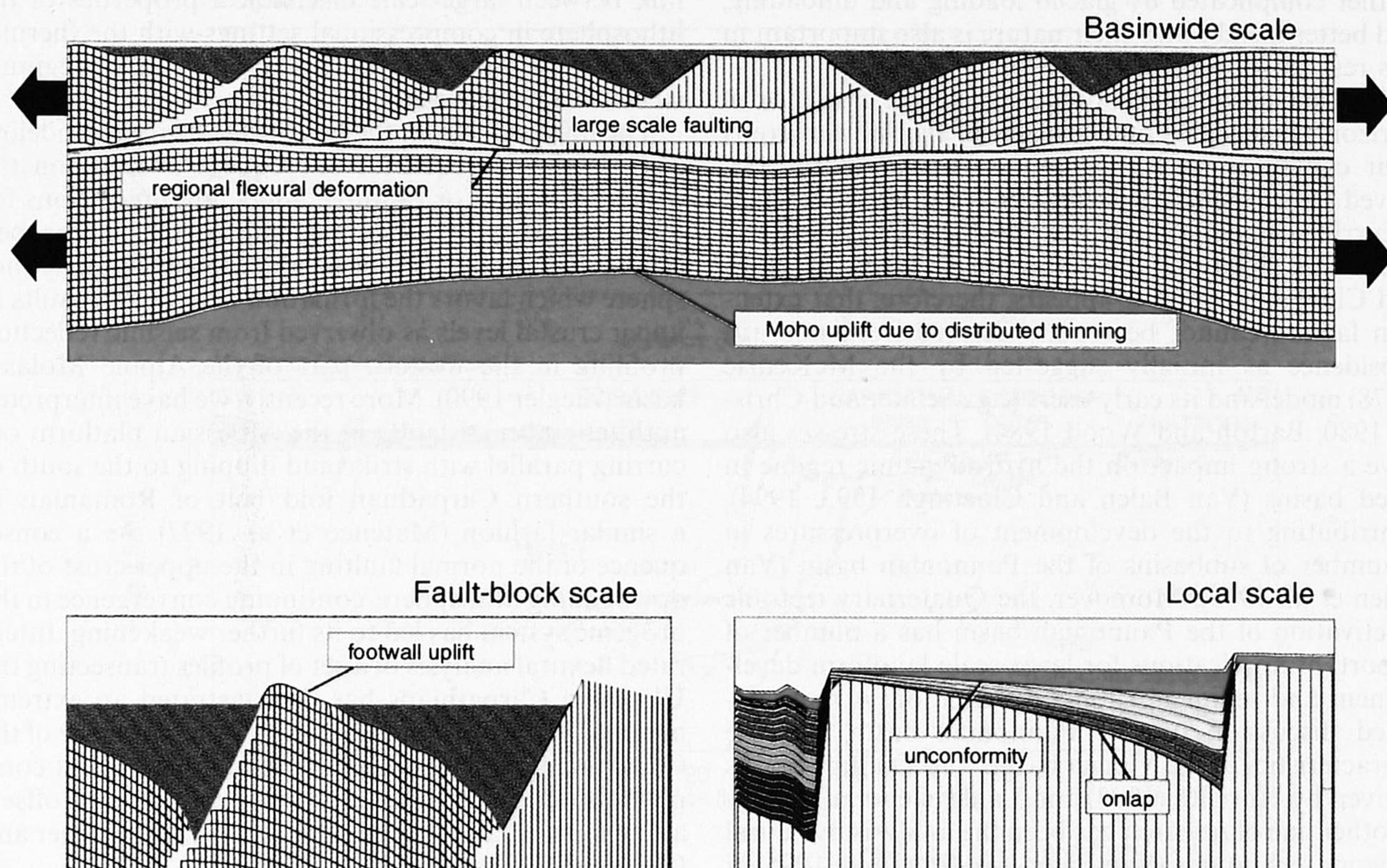
Fig. 7a, b Evolution of a hinterland basin and retreat of a rift shoulder. **b** Close-up of the stratigraphy clearly demonstrating the offlapping nature of the first post-rift sediments caused by flexural uplift in response to rift shoulder erosion (after Van Balen et al. 1995)

hinterland basins with a stratigraphic signature distinctly different from standard models invoking stretching and post-rift flexure, commonly applied in the existing packages. Testing of this model against a number of rifted margins around the world demonstrates that erosion of the rift-shoulder topography, created during extension of a lithosphere with finite strength, eliminates to a large extent the need to invoke eustatic sea-level changes to explain the most commonly encountered large-scale stratigraphic features in rifted basins and associated hinterland basins.

### Coupling of synrift faulting with large-scale flexural properties of extending lithosphere

On a subbasin scale the control by faulting on basin stratigraphy is obviously of key importance. Our modeling research has focused on the coupling of fault block tilting and the flexural behavior of the extending lithosphere. Our modeling approach, as illustrated in Fig. 8, is twofold. In a first step modeling technology was developed to quantify the thermal effects of faulting in an extending lithosphere (Ter Voorde and Bertotti 1994); this was followed by the development of models linking the evolution of individual half-grabens to deeper lithospheric responses (Ter Voorde and Cloetingh 1996). The second step was aimed at validating and testing these models in well-constrained natural laboratories. To this end we have selected for testing of the thermal signature of fault-controlled extension the well-exposed Mesozoic Southern Alpine rifted margin, where extensive radiometric dating and the application of Ar/Ar laser-probing techniques has





**Fig. 8** Finite difference model developed for tilted fault-block modeling. Large-scale deformation is coupled to fault-block and local-scale phenomena. (After Ter Voorde and Cloetingh 1996)

enabled us to constrain with high accuracy the thermal evolution of the extending lithosphere at subbasin and basinwide scales (Bertotti and Ter Voorde 1994). Testing and validation of the stratigraphic modeling component was carried out by a case-history modeling of the Oseberg field in the Norwegian part of the Viking graben (Ter Voorde et al. 1996). This study demonstrated the need to establish a regional framework, linking the mechanics of the Oseberg block to the crustal evolution of adjacent areas, as a pre-requisite for subsequent reservoir modeling.

Considering the notion that different spatial and temporal scales in synrift basins are by their very nature linked, ignorance of constraints and structural information on surrounding areas will severely limit the quality of reservoir modeling. Modeling provides a quantitative tool to estimate the tectonic control on synrift depositional sequences on a subbasin scale (see also Nottvedt et al. 1995). From our research results we conclude that the amount of footwall uplift of an individual fault block depends directly on the level of necking, controlling the large-scale response of the lithosphere to extension, and, therefore, should not

only be attributed to more local factors restricted to the subbasin scale.

#### **Tectonic control on post-rift evolution of extensional basins**

It should be noted that software packages presently available on the market generally do not have the capability to incorporate the effects of stress changes on basin architecture and basin fill. The realization that during the post-rift evolution of extensional basins the stress field can change dramatically has a number of important implications for the post-rift evolution of such basins (Ziegler et al. 1995). This notion is important, as extensional basins, such as the North Sea, the Gulf of Lion, and the Pannonian basin, are presently in a state of horizontal compression as documented by stress indicators summarized in the World Stress Map (Zoback 1992) and the European Stress Map compiled in Karlsruhe (e.g., Gölke et al. 1996). Stresses in the lithosphere affect its vertical motions during the post-rift phase, causing tilting of the basin margins with a magnitude dependent on the ratio of the level of the stresses and the strength of the lithosphere inherited from the synrift phase. Horizontal stresses in the lithosphere also strongly affect the post-rift stratigraphy and the development of diapirism in rifted basins (Cloetingh and Kooi 1992). Post-rift lithospheric motions are



further complicated by glacial loading and unloading, and better insight into their nature is also important in this respect (Solheim et al. 1996).

Another important consequence of the presence of horizontal compressional stresses in the lithosphere is their capability to contribute to the commonly observed Late Cenozoic accelerations in basin subsidence, occurring in the North Sea basin (Van Wees and Cloetingh 1996) and the Pannonian basin (Horváth and Cloetingh 1996). It appears, therefore, that extension factors cannot be readily derived from post-rift subsidence as initially suggested by the McKenzie (1978) model and its early users (e.g., Sclater and Christie 1980; Barton and Wood 1984). These stresses also have a strong impact on the hydrodynamic regime in rifted basins (Van Balen and Cloetingh 1993, 1994), contributing to the development of overpressures in a number of subbasins of the Pannonian basin (Van Balen et al. 1997). Moreover, the Quaternary tectonic reactivation of the Pannonian basin has a number of important implications for large-scale landform development and sediment supply distribution. A more detailed discussion on the Pannonian basin and the characteristics of the Vienna basins and Danube basins is given by Horváth (1993) and Lankreijer et al. (1995). Another line of research has been the analysis by neural network technology (Van Balen and Cloetingh 1995) of the effect of stress-induced overpressures, linking backward analysis to predictions of forward modeling. The effects of intraplate stresses on the evolution of the basins on the Russian platform forms also a research topic carried out in the framework of the International Lithosphere Program (ILP) EUROPROBE project (Cloetingh and Lobkovsky 1996; Lobkovsky et al. 1996; Nikishin et al. 1996).

### Compressional basins

Lateral variations in flexural behavior and implications for paleotopography

Our modeling of compressional basins located in the Alpine belt of Europe followed the same philosophy as taken in the extensional basin modeling approach. We started with modeling on a lithosphere scale, with a focus on the role of flexural behavior of the lithosphere in foreland basin evolution (e.g., Zoetemeijer et al. 1990; Van der Beek and Cloetingh 1992). These studies could build on data sets obtained from wells, gravity data, and deep seismic profiling, such as the ECORS profile through the Pyrenees completed in the last 1980s. A detailed discussion on constraints provided by deep seismic data on the bulk geometry of Alpine belts is given by Roure *et al.* (1996a) and Ziegler and Roure (1996). Flexural modeling studies were backed up by modeling of the large-scale continental rheology evolution (Cloetingh and Burov 1996), demonstrating a direct

link between large-scale mechanical properties of the lithosphere in compressional settings with the thermal structure of the lithosphere and the level of regional stresses in the lithosphere.

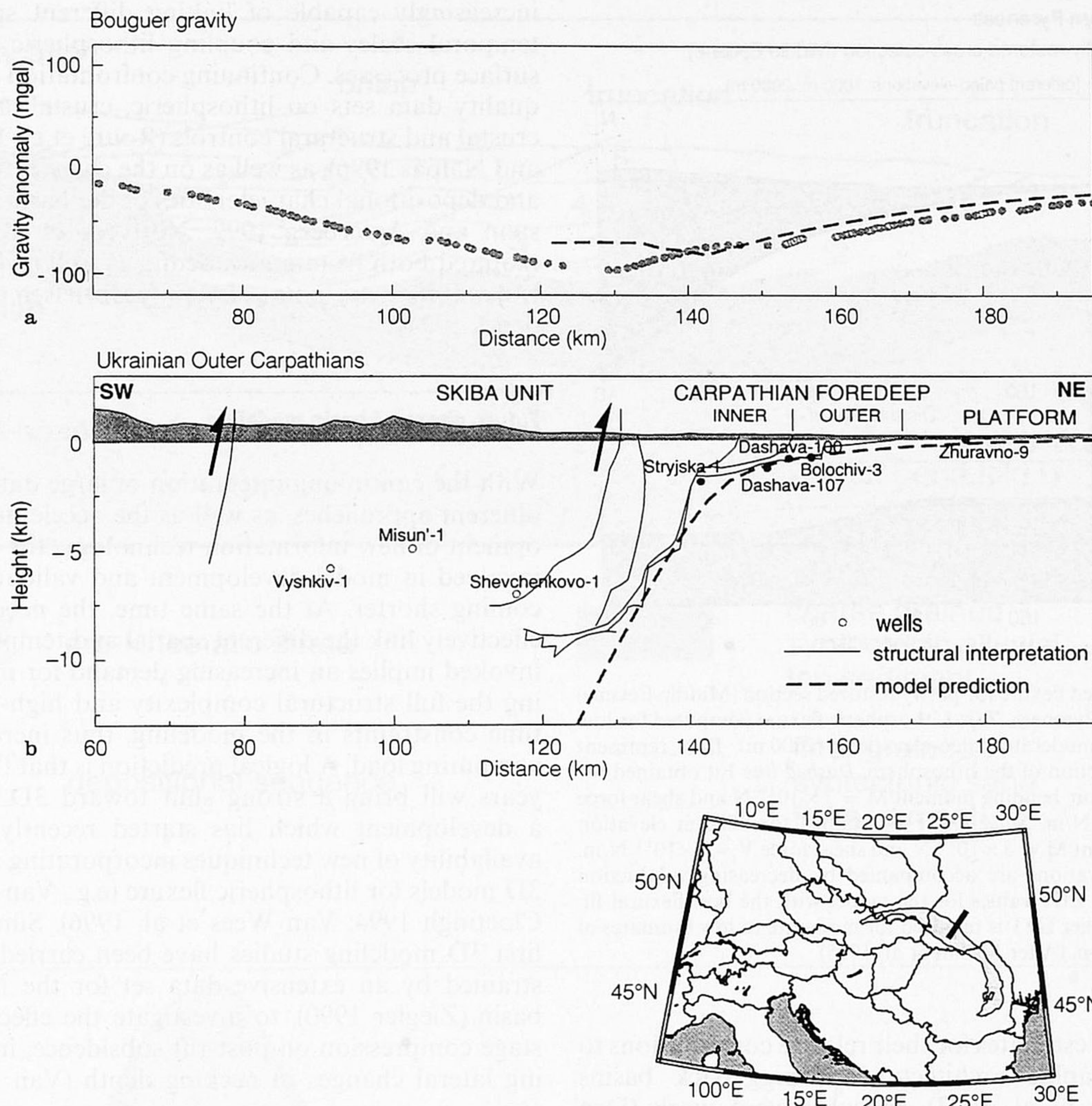
The inferences from the large-scale flexural modeling provide the feedback for subsequent analysis on the subbasin scale. For example, modeling predictions for the presence of weak lithosphere in the Alpine belt invoke the occurrence of steeply downbending lithosphere which favors the formation of tensional faults at upper crustal levels as observed from seismic reflection profiling in the western part of the Alpine Molasse basin (Ziegler 1990). More recently, we have interpreted antithetic normal faults in the Moesian platform occurring parallel with strike and dipping to the south of the southern Carpathian fold belt of Romania in a similar fashion (Matenco et al. 1997). As a consequence of the normal faulting in the upper crust of the downbending lithosphere, continuing convergence in the orogenic system has led to its further weakening. Integrated flexural analysis of a set of profiles transecting the Ukrainian Carpathians has demonstrated an extreme bending of the lithosphere in the Ukrainian sector of the Carpathian system (see Fig. 9), leading to almost complete lithospheric failure and very pronounced offsets along normal faults in the upper crust (Zoetemeijer and Cloetingh 1997).

After a number of studies of the paleo-rheology of the lithosphere constrained by high-quality thermochronology in the central Alps (Okaya et al. 1996) and eastern Alps (Genser et al. 1996), the importance of the presence of large lateral variations in mechanical structure in mountain belts became evident. In particular, it was realized that a pronounced reduction in strength generally occurs from the external part of the orogen towards its inner parts. The consequences of the lateral variations in flexural strength of the lithosphere were explored in a modeling study carried out along a transect through the NE Pyrenees, which is well constrained by deep seismic crustal control as well as an extensive field-derived database (Verges et al. 1995). Apart from investigating the present basin configuration, and quantifying the present-day mechanical structure underlying the southern Pyrenees fold-and-thrust belt, we also investigated the relationships between paleo-topography and flexural evolution of the orogen (Millan et al. 1995). This novel approach has led to a set of predictions on paleotopography, with important implications for sediment sourcing in the area (Fig. 10). Predictions of these models will be tested and integrated with estimates on paleo-topography and uplift histories derived from fission-track studies.

Controls of thrusting on foreland basin stratigraphy

On a smaller scale, the architecture of the proximal parts of foreland basins is strongly affected by thrusting.





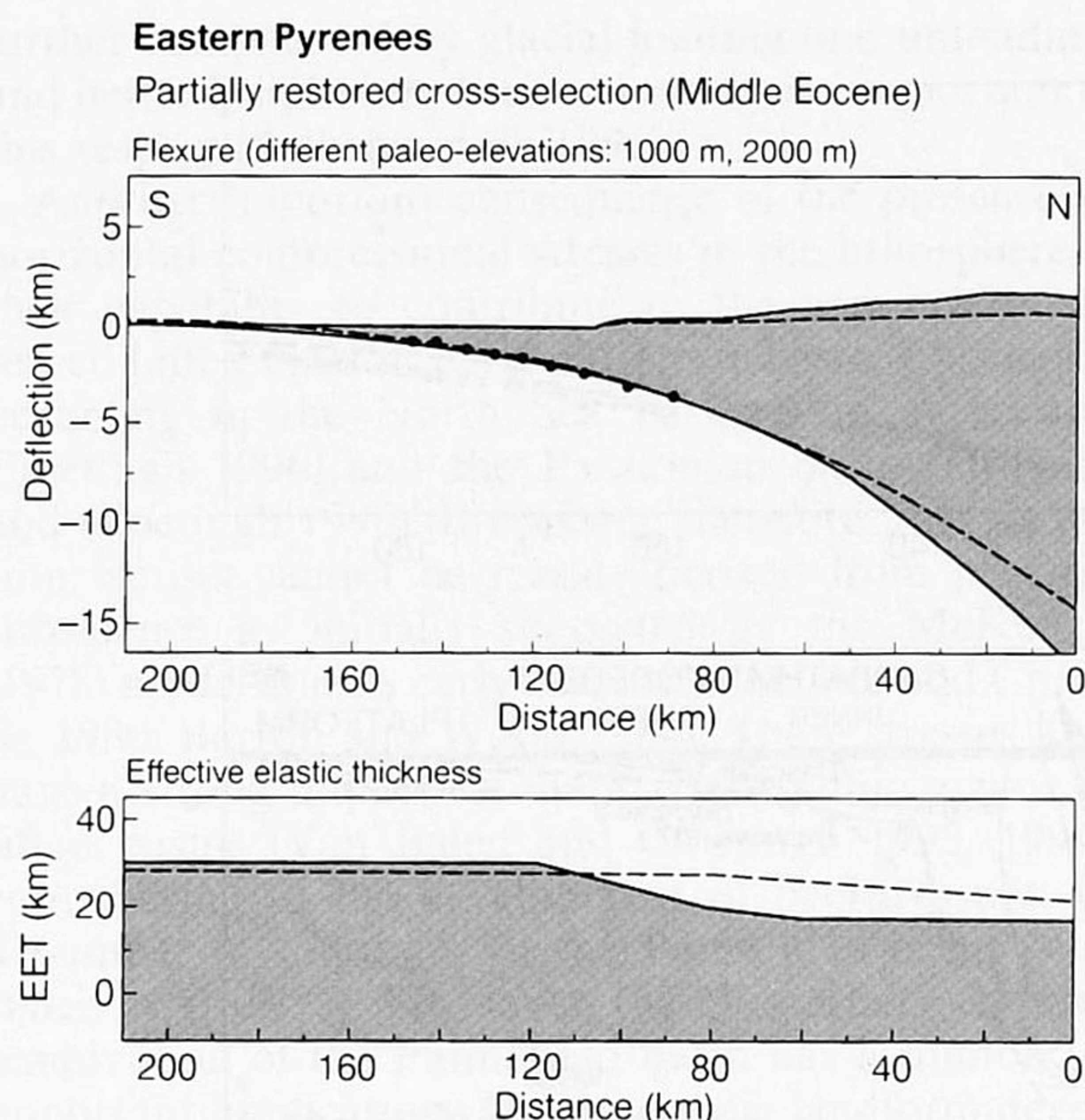
**Fig. 9a, b** Cross section illustrating the steep downbending of the East European platform lithosphere under the Ukrainian segment of the Carpathian foreland fold-and-thrust belt system. The extreme curvature of the downbending lithosphere associated with mechanical weakening caused by plate-boundary processes operating along the Carpathian arc system leads to extensive normal faulting in the upper crust. Characteristic for this situation where foreland flexure develops on weak lithosphere is the presence of a narrow foreland basin. (After Zoetemeijer and Cloetingh 1997)

Detailed studies of the basin fill architecture and the field relationships in the southern Pyrenean foreland fold-and-thrust belt basins have demonstrated the importance of a tectonic control on the sedimentary record (e.g., see Puigdefabregas et al. 1992). Our modeling approach focused on two lines. One line of research has concentrated on the effects of changes in accommodation shape induced by stress fields and orogenic wedge growth, occurring as a result of thrusting, and their effects on the large-scale sedimentary record in the

distal parts of the foreland fold-and-thrust belts (Peper et al. 1994). We have explored the consequences of these models for the southern Pyrenean fold-and-thrust belt to investigate the effects of thrusting-related "cycles" in the context of climate-driven and autocyclic components in the sedimentary record (Peper and de Boer 1995; Peper and Cloetingh 1995).

In a complementary approach, we investigated the direct consequences of thrusting on the stratigraphy in the proximal basins in foreland fold-and-thrust belt systems. This work was carried out to address the need for better constraints on the time control on the sequence of thrusting in the outer part of foreland fold-and-thrust-belt systems. To this aim, forward kinematic modeling based on the principles of balancing techniques was developed incorporating syntectonic sedimentation and lithospheric flexure (Zoetemeijer et al. 1992). This approach enabled the quantification of the effects of simultaneously occurring tectonic processes and eustatic sea-level changes, leading to





**Fig. 10** Modeled flexure for partly restored section (Middle Eocene) in the eastern Pyrenees. *Top*: Lithospheric flexure computed for low (1000 m) and moderate paleo-elevations (2000 m). *Dots* represent observed deflection of the lithosphere. *Dashed line* Fit obtained for 1000-m elevation, bending moment  $M = 7 \times 10^{16}$  N and shear force  $V = 1.6 \times 10^{11}$  N/m. *Solid line* Fit obtained for 2000-m elevation bending moment  $M = 5 \times 10^{16}$  N and shear force  $V = 1 \times 10^{11}$  N/m. Increasing elevations are accompanied by decreasing subduction forces. *Bottom*: EET values for the model with the best flexural fit. An overall thicker EET is required for moderate to low estimates of palaeo-elevation. (After Millan et al. 1995)

quantitative estimates for their relative contributions to the stratigraphic architecture of piggyback basins (Zoetemeijer et al. 1993). In subsequent work (Den Bezemer et al. 1997) we modeled the consequences of different modes of thrusting and erosion on the stratigraphic record in compressional settings in the vicinity of the thrust front (see Fig. 11). These models allow the full exploration of different scenarios for fault activation, sense of thrusting, depth of fault activation, and spacing of thrusts. This approach has the capability to integrate in an adequate way constraints provided by structural geology with basin modeling in compressional settings. Such constraints can result from analog modeling (see Brun and Nalpas 1996), computer modeling (see Sassi et al. 1993; Zoetemeijer et al. 1993; Roure and Sassi 1995; Toth et al. 1996), seismic reflection data, and field studies in onshore and offshore basins (see Roure et al. 1996b). These models can also be linked with the large-scale flexural models discussed above. This approach provides a major step forward, particularly when combined with available cross-section balancing packages. This research has also set the stage for a coupling with facies modeling (Walker and James 1990) which will be an important target for continued future modeling activity. Basin modeling is

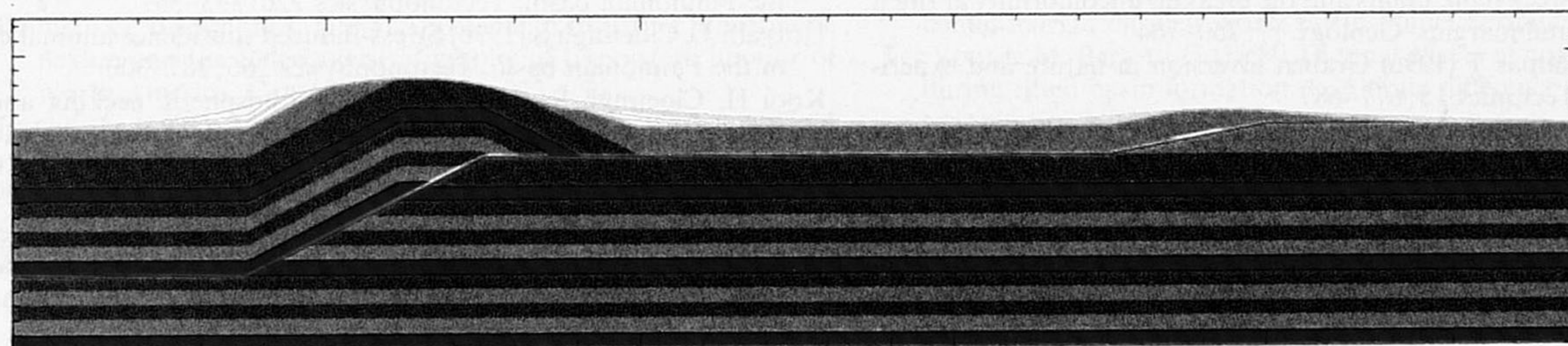
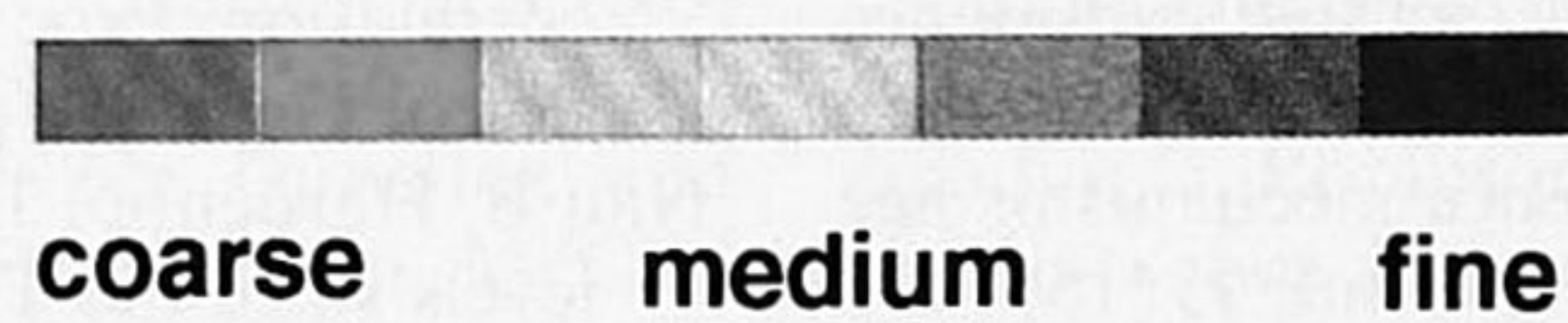
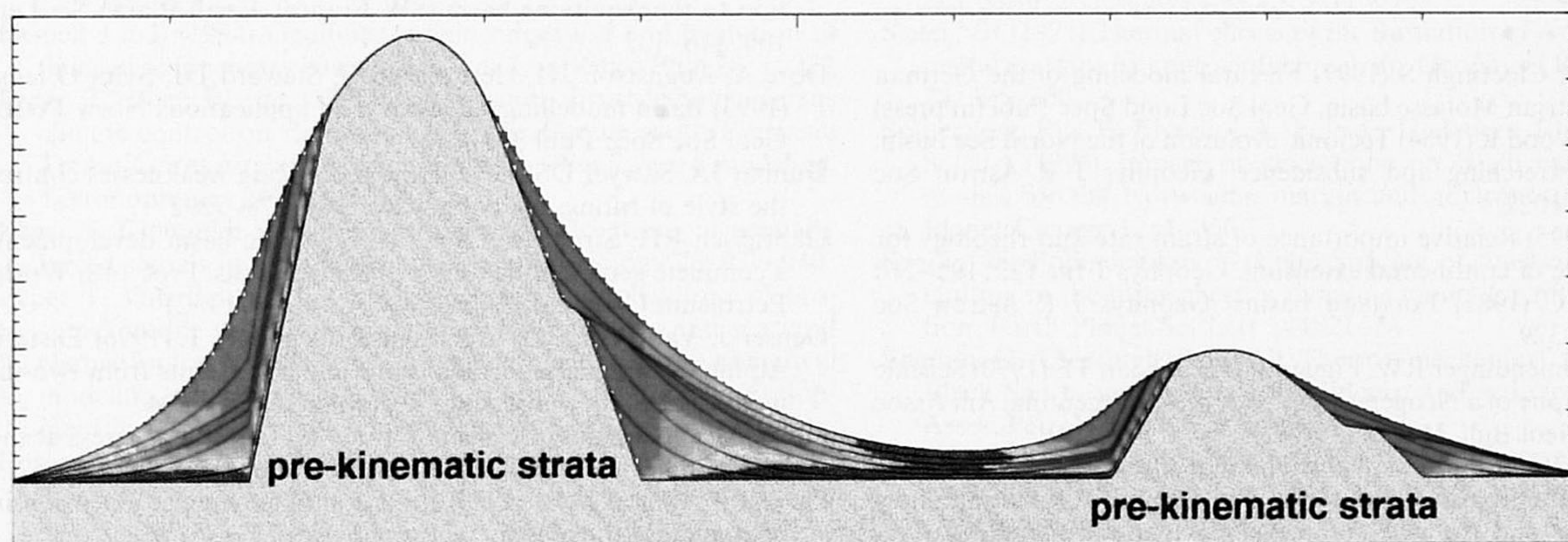
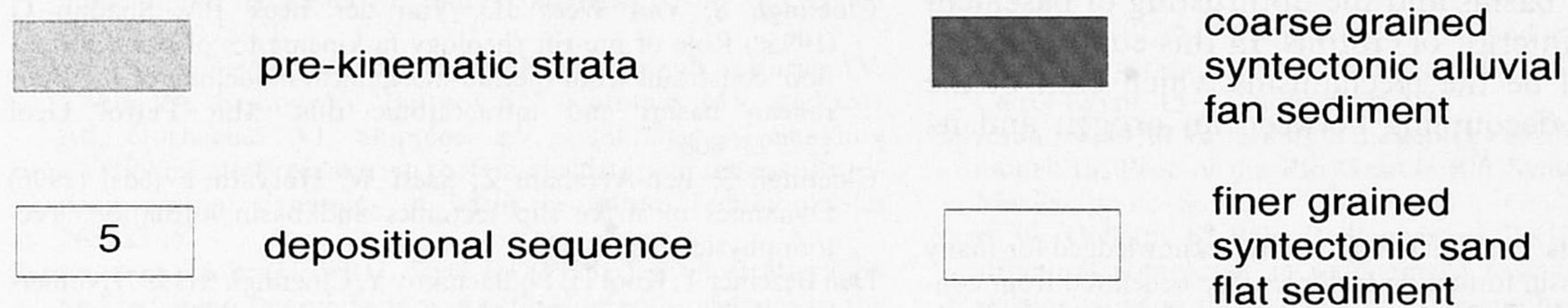
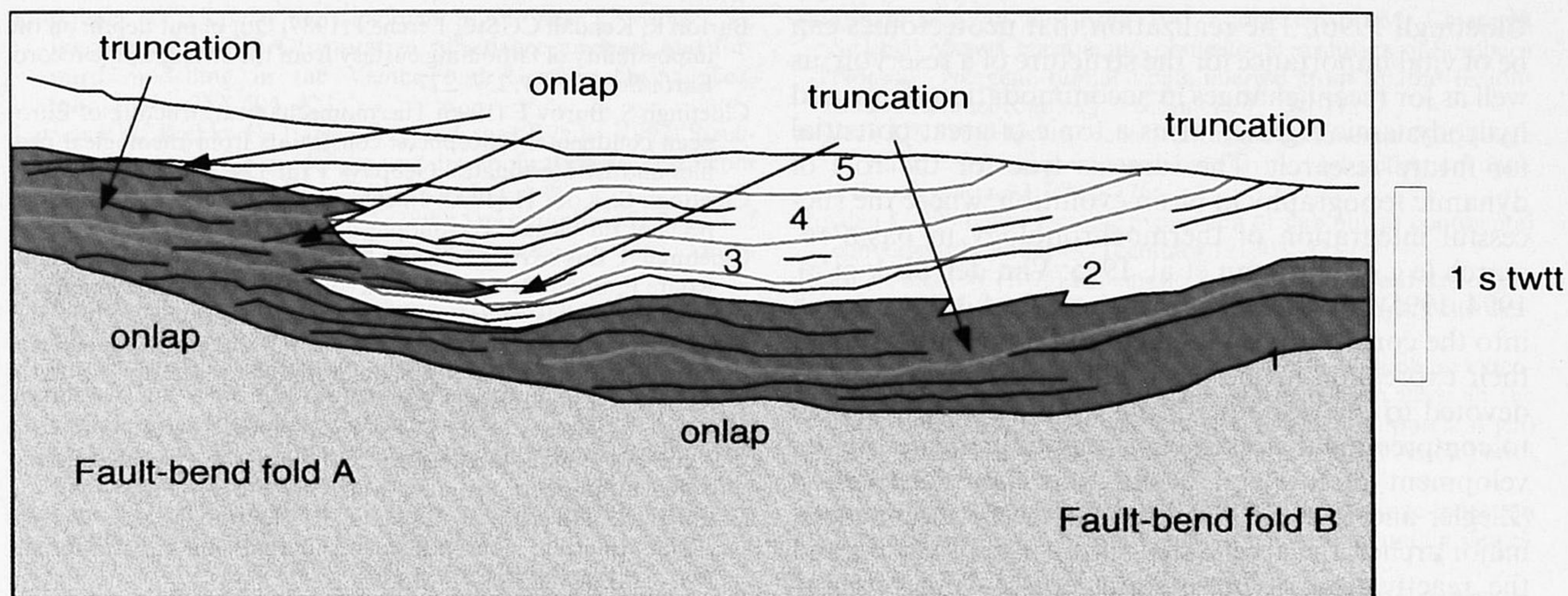
increasingly capable of linking different spatial and temporal scales and coupling lithospheric and near-surface processes. Continuing confrontation with high-quality data sets on lithospheric, crustal, and supra-crustal and structural controls (Roure et al. 1994; Brun and Nalpas 1996), as well as on the chronostratigraphy and depositional characteristics of the basin fill (Gradstein and Agterberg 1997; Nottvedt et al. 1995), is required both from an academic as well as from a petroleum industry point of view (Gabrielsen and Strandes 1994).

### Future phase of basin modeling

With the continuing integration of large data sets and different approaches, as well as the accelerating development of new information technology, the cycle time involved in model development and validation is becoming shorter. At the same time, the need to more effectively link the different spatial and temporal scales invoked implies an increasing demand for incorporating the full structural complexity and high-resolution time constraints in the modeling, thus increasing the computing load. A logical prediction is that the coming years will bring a strong shift toward 3D modeling, a development which has started recently with the availability of new techniques incorporating faulting in 3D models for lithospheric flexure (e.g., Van Wees and Cloetingh 1994; Van Wees et al. 1996). Similarly, the first 3D modeling studies have been carried out, constrained by an extensive data set for the North Sea basin (Ziegler 1990), to investigate the effects of late-stage compression on post-rift subsidence, incorporating lateral changes in necking depth (Van Wees and

**Fig. 11** Comparison of modeling of compressional basins on sub-basin scale with interpreted seismic section through typical frontal part of an orogenic system. *Top*: Interpreted seismic section of the Iglesia basin (Argentina) showing sediments deposited in an intra-montane basin. The basin is thought to be underlain by two coupled fault-bend folds which governed sedimentation. The syntectonic depositional sequence shows an alternating pattern of truncation and onlap with associated interfingering of coarse and finer facies. (After Beer et al. 1990). *Bottom*: Simulation of sedimentary infill (magnified in top part) associated with coupled fault-bend fold pair (shown in bottom part). The central basin located between the two folds corresponds to the interpreted seismic section displayed in the top panel. Onlap and truncation patterns predicted by the model are caused by fault displacement velocity variations as suggested by Beer et al. (1990). Similarly, as anticipated by the interpreters of the seismic section, the prograding of coarse sediments occurs simultaneously with truncation, whereas retrograding of coarse sediments coincides with phases of onlap. In contrast to the facies interpretation favored by Beer et al. (1990), the modeling predicts separation of coarse sediment bodies attached to both folds developed in the basin. The modeling also suggests an extrabasinal control to explain the observed fining-upward sequence (see upper panel), as in the absence of such a control the model predicts a general coarsening-upward sequence. (After Den Bezemer et al. 1997)







Cloetingh 1996). The realization that neotectonics can be of vital importance for the structure of a reservoir, as well as for recent changes in accommodation shape and hydrodynamic regime, forms a topic of great potential for future research. The same is true for the role of dynamic topography in basin evolution, where the successful integration of thermochronology in basin research (e.g., Rohrmann et al. 1995; Van der Beek et al. 1994, 1995) has opened new avenues for future research into the constraints on different uplift mechanisms and their expression in the basin fill. Major efforts will be devoted to the response of the continental lithosphere to compressional stresses, not only influencing the development of foreland basins and their destruction (Ziegler and Roure 1996), but also the development of major arches, the accelerated subsidence of basins, and the reactivation of pre-existing crustal and possibly lithospheric weakness zones, controlling the inversion of extensional basins and the upthrusting of basement blocks in the interior of cratons. In this context a crucial point will be the mechanisms which control the coupling and decoupling between an orogen and its foreland plate.

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